GEOPHYSICAL RESEARCH LETTERS, VOL. 19, NO. 2, PAGES 215-218, JANUARY 24, 1992

POTENTIAL CLIMATE IMPACT OF MOUNT PINATUBO ERUPTION

James Hansen, Andrew Lacis, Reto Ruedy and Makiko Sato

NASA Goddard Space Flight Center Goddard Institute for Space Studies, New York

Abstract. We use the GISS global climate model to make a preliminary estimate of Mount Pinatubo's climate impact. Assuming the aerosol optical depth is nearly twice as great as for the 1982 El Chichon eruption, the model forecasts a dramatic but temporary break in recent global warming trends. The simulations indicate that Pinatubo occurred too late in the year to prevent 1991 from becoming one of the warmest years in instrumental records, but intense aerosol cooling is predicted to begin late in 1991 and to maximize late in 1992. The predicted cooling is sufficiently large that by mid 1992 it should even overwhelm global warming associated with an El Nino that appears to be developing, but the El Nino could shift the time of minimum global temperature into 1993. The model predicts a return to record warm levels in the later 1990s. We estimate the effect of the predicted global cooling on such practical matters as the severity of the coming Soviet winter and the dates of cherry blossoming next spring, and discuss caveats which must accompany these preliminary simulations.

Introduction

On June 15, 1991, nature launched her own great climate experiment, as the explosion of the Philippine volcano Mt. Pinatubo sent massive amounts of gas and dust to heights of more than 25 km. The global shield of stratospheric aerosols caused by Pinatubo will probably have an opacity that exceeds any volcano of the past century, thus producing a climate forcing large enough to provide a valuable test of global climate models. Volcanic aerosols reflect sunlight to space and thus reduce solar heating of the Earth, a process recognized by Benjamin Franklin (1784) who argued that a "dry fog" covering the Northern Hemisphere in 1783-1784, arising from an Icelandic volcano, was probably the cause of unusually cold weather at that time. Such speculation was further fueled by two huge 19th century volcanos: Tambora in 1815, which was followed by "the year without a summer" in the United States [Stommel and Stommel, 1983], and Krakatau in 1883, which was followed by the coolest year (1884) recorded during the period (1880-present) of near-global thermometer measurements [Hansen and Lebedeff, 1987].

Quantitative analyses defining the climate impact of volcanos are lacking, because of inadequate data on aerosol radiative forcing and because the large natural variability of climate tends to mask any volcanic signature [Robock, 1991; Self and Rampino, 1988; Mass and Schneider, 1977]. The first large volcano with measurements of aerosol microphysical properties and their global distribution was the 1963 Mt. Agung eruption. Calculations for Agung with a one-dimensional radiative-convective climate model [Hansen et al., 1978] yielded surface cooling of a few tenths of a degree Celsius and stratospheric warming of a few degrees, consistent with observations; but the aerosol data were crude, the climate model was very simple with somewhat arbitrary parameters, and the temperature changes were only of the order of interannual variability. The one subsequent large volcanic climate perturbation, El Chichon in 1982, was complicated by a nearly simultaneous El Nino, and again any global climate effects appeared to be near the level of natural climate variability [Angell and Korshover, 1984]. Thus the possibility of an even larger and more precisely observed volcanic radiative perturbation is of great interest.

In this paper we show that Pinatubo aerosols should provide an acid test of climate models. Although simulated global coolings for El Chichon and Agung were at about the 1½ σ level, where σ is the standard deviation of annual global mean temperature, Pinatubo appears to be at the 3 σ level. Below we define our assumptions about the Pinatubo aerosols, describe the global climate model simulations, and discuss a few implications of the predicted aerosol cooling.

Paper number 91GL02788
0094-8534/92/91GL-02788$03.00

Aerosol Climate Forcing

Aerosol radiative forcing of the climate system depends upon the aerosol geographical distribution, optical depth, size distribution, composition and altitude. However, as we show in another paper [Lacis et al., 1992, hereafter L92], stratos-pheric aerosol radiative forcing of tropospheric climate is primarily a function of the aerosol column optical depth, τ. Also, the effective radius of the aerosol size distribution is important: if there are a substantial number of particles larger than 1 μm, a condition which may occur during the first few months after a volcanic eruption, the aerosol cooling effect is diminished.

We perform GCM simulations with three assumptions for aerosol opacity, labeled El, 2•El and P. El has Pinatubo aerosol properties identical to those in our earlier simulations of El Chichon [Hansen et al., 1988, hereafter H88]. The aerosols are a 75% solution of sulfuric acid in water, with sizes based on the May and October distributions of Hofmann and Rosen (1983); the fraction of optical depth based on the May distribution decreases linearly from 1 at the time of eruption to 0 six months later. The aerosols initially were distributed uniformly between the equator and 30 N, later spreading globally with τ twice as large at 30-90 N as at 30 N to 90 S. Beginning 10 months after the eruption, τ decayed exponentially with a 12-month time constant. Although this scenario was based on early El Chichon data, the time dependence of global optical depth is similar in a later analysis by J. Pollack (private communication, 1985), with our optical depth about 10% larger than Pollack's. In the 2•El experiment τ is twice as large as in the El experiment, in recognition of early reports that sulfur emissions from Pinatubo may have been twice as large as for El Chichon [A. Krueger, private communication].

The P experiments have the same time dependence of global optical depth as the El and 2•El experiments, but with τ 1.7 times larger than in El and the aerosol geographical distribution modified as described below. These changes crudely account for information on Pinatubo provided at an interagency meeting in Washington D.C. on September 11 organized by Lou Walter and Miriam Baltuck of NASA, including aerosol optical depths estimated by Larry Stowe from satellite imagery. The global τ and radiative forcing at the tropopause, ΔF (Figure 1), are uncertain by perhaps 50%, but we believe ΔF is conservatively estimated. ΔF does not increase as rapidly as τ during the first six months, because of the changing fraction of large particles during that period (paper L92). In scenario P aerosols are restricted uniformly to latitudes 20 S to 30 N in the first three months after the eruption, after which these hemispheric amounts spread poleward into both hemispheres, becoming hemispherically uniform in January 1992. The geographical distribution of the forcing is also affected by the aerosol scattering approximation in the GCM, which overestimates aerosol radiative forcing at low latitudes and underestimates it at high latitudes. These uncertainties should not greatly impact the semi-quantitative conclusions of this paper.

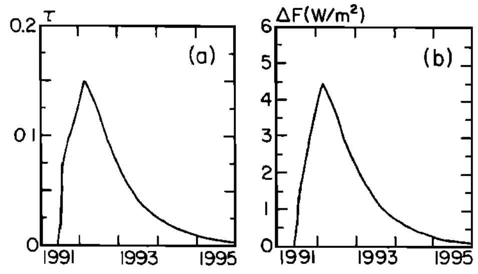

Fig. 1. Global mean aerosol optical depth and radiative forcing at tropopause for Pinatubo (P_n) simulations.

We use two control runs for these experiments: scenarios A and B of paper H88. Both began in 1958 and include climate forcing due to increasing greenhouse gases and changing stratospheric aerosols. Scenario A has fast (exponential) growth rates for greenhouse gases and no volcanic aerosols after 1985. Scenario B has linear growth of greenhouse gases and an El Chichon sized volcano in 1995. Scenarios El, 2•El and P do not include additional volcanos in the 1990s after Pinatubo.

Figure 2 shows the effect of El and 2•El aerosols on simulated global mean temperature. Aerosol cooling is too small to prevent 1991 from being one of the warmest years this century, because of the small initial forcing and the thermal inertia of the climate system. However, dramatic cooling occurs by 1992, about 0.5°C in the 2•El case. The latter cooling is about 3 σ, where σ is the interannual standard deviation of observed global annual-mean temperature. This contrasts with the 1½ σ coolings computed for the Agung (1963) and El Chichon (1982) volcanos. If the 2•El aerosol amount is realistic, the predicted cooling far exceeds *global* warmings associated with El Ninos, which are typically about 0.2°C [Angell, 1988; Hansen and Lebedeff, 1988; Jones, 1989].

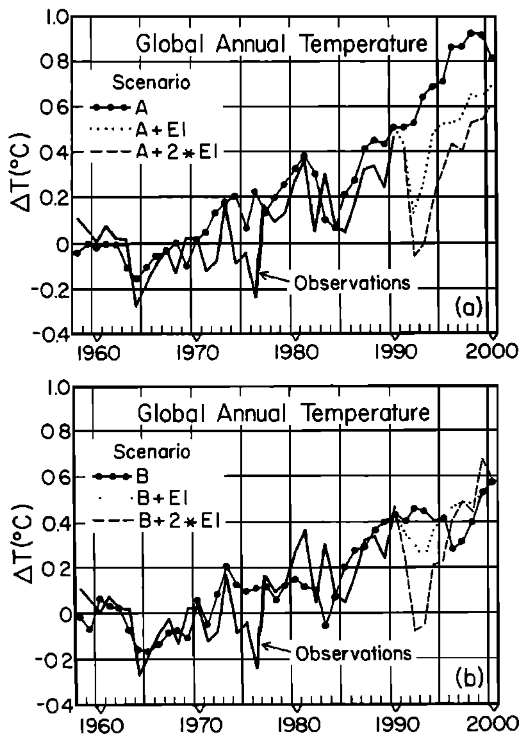

Fig. 2. Annual-mean global surface air temperature computed for scenarios A, A+El, and A+2•El (a) and B, B+El, and B+2•El (b). Observational data are an update of Hansen and Lebedeff (1987). Zero point for observations is 1951-1980 mean; model zero point is 100 year control run mean.

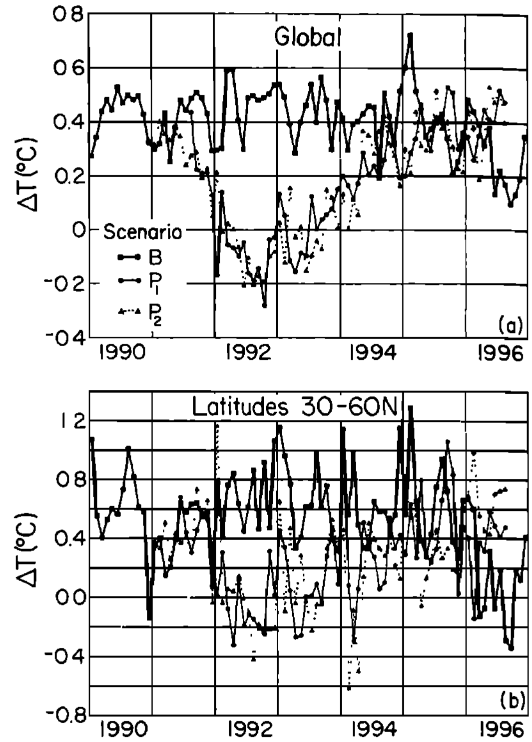

Fig. 3. Monthly mean global (a) and northern latitude (b) surface air temperature for scenarios B, P_1 and P_2.

The predicted global cooling is examined in finer temporal and spatial detail for scenario P, which was run twice. The P_1 simulation began June 1, 1991 with scenario B initial conditions. P_2 had the same climate forcing as P_1, but it began with scenario B initial conditions on December 1, 1990, to which random changes of atmospheric temperature up to 1°C were added. Thus P_2 provides a "noise level" comparison, illustrating how natural variability impacts our ability to discern a volcanic signature in global and regional temperatures.

Monthly temperatures for scenarios B, P_1 and P_2 are shown in Figure 3. The global temperature in the P scenarios declines rapidly, reaching a peak cooling of about 0.5°C in late 1992. The cooling is greater at mid northern latitudes (Figure 3b), but less obvious there because of the larger natural variability.

The volcano "signal" is harder to detect as the area examined becomes smaller, especially at high latitudes where atmospheric variability is large. Furthermore, real world variability appears to be larger than the variability in our model, as shown, for example, by comparisons with monthly, seasonal and annual observations [Hansen and Lebedeff, 1988]. The lesser variability in the model may be due to the absence of variable ocean dynamics in our simple treatment of ocean heat transport (H88). This is especially evident at low latitudes, where observations show large El Nino warming events at about 3-7 year intervals.

A graphic illustration of signal and noise is provided by maps of seasonal mean temperature anomalies for the first six seasons after the Pinatubo eruption, calculated for scenarios B, P_1 and P_2 (Figure 4). In the Pinatubo experiments a cooling tendency appears at low latitudes in late 1991 and by mid 1992 the cooling is essentially global. If an El Nino begins in late 1991 as currently projected [M. Cane, private communication], much of the low latitude cooling may be offset during 1992.

Although the simulated global cooling is dramatic (Figures 3 and 4), even at the time of peak cooling we can not predict absolutely whether the seasonal mean temperature at a given locale will be above or below "normal". For example, for a given region, such as the United States, scenarios P_1 and P_2 often show the opposite sign for the seasonal mean temperature

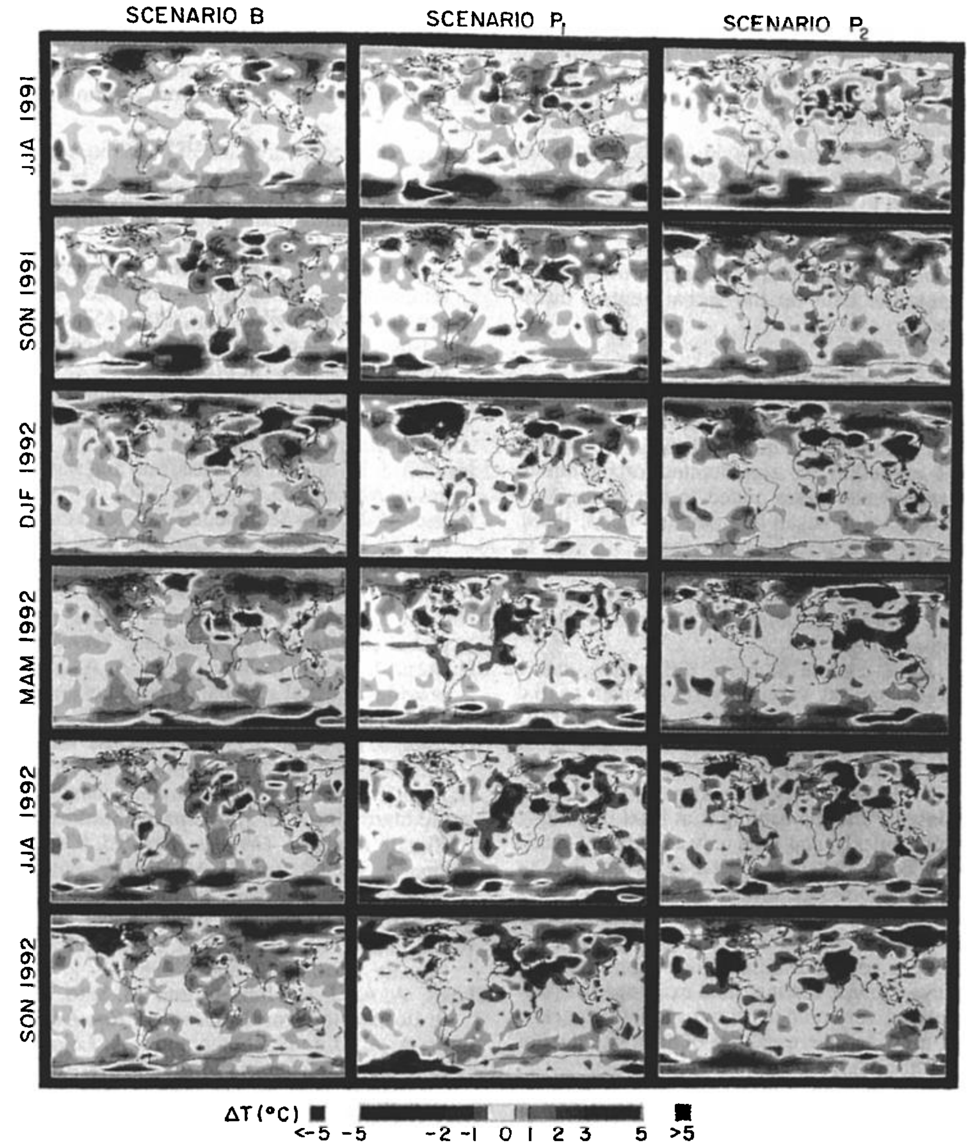

Fig. 4. Simulated seasonal surface temperature anomalies, relative to 100 year control run with 1958 atmospheric composition, in scenarios B, P_1 and P_2 for first six seasons after Pinatubo eruption.

anomaly (Figure 4), even though the two scenarios have identical climate forcings. This is expected climatic behavior since the standard deviation of local seasonal mean temperature, σ, is generally larger than the simulated 0.5°C global cooling. At midlatitudes the observed σ is about 0.5-1°C for summer and 1-2°C for winter [Hansen and Lebedeff, 1987]. Nevertheless, the simulated aerosol cooling is large enough to alter noticeably the probability of a warmer than normal or cooler than normal season, as is obvious from comparison of scenario B with P_1 and P_2 in Figure 4.

Is this predicted aerosol cooling large enough to be noticeable to the man-in-the-street? We plan to examine this issue in detail by running several dozen Pinatubo scenarios with slightly altered initial atmospheric conditions and using these for statistical studies. In the interim, we use the computed zonal mean temperature anomalies in scenarios P_1 and P_2 to estimate the expected seasonal mean surface air temperature anomaly for a given locale. This anomaly is added uniformly to a 30 year (1950-1979) climatology for that locale, as in H88. We define the 10 warmest summers in the 30 year climatology as hot, the next 10 as normal or average, and the remaining 10 as cold, and similarly for the other seasons. Thus, in the absence of a long term trend, the probability of an unusually hot, a normal, and an unusually cold season are each about 33%.

With these definitions, the probability of an unusually cold winter in Moscow decreases from 33% in the 1950s to 15-20% in recent years in scenario B, as a result of greenhouse warming. But the Pinatubo aerosols increase this probability to 30-50% in each of the three winters 1991-2, 1992-3 and 1993-4. Thus the model predicts a substantial increase in the likelihood of a severe winter. We made a similar estimate for the change of the probability of a hot summer in Omaha, in the corn belt of the United States. In scenario B this probability reaches 60% in the early 1990s, but with the Pinatubo aerosols it declines to about 30% in 1992 and 1993, returning to the 60% level by 1996.

Another measure of climate change which might be noticeable to the man-in-the-street is the date of cherry blossoming in Washington and Tokyo. In both cities the climatological seasonal warming rate is about 1°C per week. The simulated early spring warming between the 1950s and the early 1990s in scenario B is enough to shift the time of maximum blossoming earlier by nearly a week. In the P scenarios the cooling of midlatitude land regions in the Northern Hemisphere springs of

1992, 1993 and 1994 is sufficient to shift the time of maximum blossoming about one week later than in scenario B. But natural variability of the date of maximum blossoming can exceed a week, so people will not necessarily notice the change of probability of this once a year event over the 2-3 year period of Pinatubo's maximum influence.

These estimated cooling impacts at mid and high latitudes are probably conservative for two reasons: (1) the aerosol scattering approximation in the GCM underestimates mid and high latitude radiative forcing and exaggerates low latitude forcing, and (2) the 1.7•El assumption is near the low end of present aerosol estimates. Both factors will be improved in new simulations when satellite data on the global aerosol distribution are available.

Discussion

The estimated global mean climate forcing by Pinatubo aerosols is very large. In our P scenarios it peaks at about 4 W/m^2 in early 1992, exceeding the accumulated forcing due to all anthropogenic greenhouse gases added to the atmosphere since the industrial revolution began [Hansen and Lacis, 1990]. If further observations of the aerosols prove these estimates to be realistic, this volcano will provide an acid test for global climate models. The simulated global cooling in our climate model is about three times larger than the standard deviation of global mean temperature.

The impact on local climate is more difficult to detect because of the larger natural variability on smaller spatial scales. The global warmth attained in 1990-91 was about 0.4°C relative to 1950-1980, arguably at a level that could just about begin to be detected by the man-in-the-street who is willing to look for changes in the frequency of warmer than normal seasons. The simulated Pinatubo cooling is at least that large, but occurs much more quickly, and lasts only a couple of years. Its main impact may be to delay by several years the time at which global warming becomes generally obvious.

Many caveats which must accompany these climate simulations are discussed in H88. We note here several particularly relevant points. First, a key mechanism which could limit the response to a negative climate forcing, heat exchange with the deep ocean, is simulated very crudely in our model. Conceivably surface cooling may increase exchange of heat with deeper ocean layers to a greater degree than simulated with our ocean diffusion parameterization, thus limiting the surface cooling. Second, because the simulations begin in 1958, the effect of greenhouse gases added to the atmosphere earlier this century is only partially included. Because of the thermal inertia of the climate system there is some as yet unrealized greenhouse warming which may partly offset aerosol cooling [Hansen et al., 1985]. This needs to be investigated with climate simulations which begin in the 1800s. Third, the Pinatubo climate forcing depends on aerosol properties, mainly particle size, for which we do not yet have adequate data (paper L92). Also Pinatubo may have injected water vapor into the stratosphere and altered stratospheric ozone, but observational data are inadequate to define these climate forcings. Fourth, measurements of independent competing short-term and long-term climate forcings, especially changes of ozone profiles [Lacis et al., 1990] and tropospheric aerosols [Charlson et al., 1991] are grossly inadequate. Fifth, climate feedbacks, such as changes of cloud properties and atmospheric water vapor, may not be accurately simulated in our climate model. Indeed, nature's Pinatubo climate experiment provides a chance to check climate feedback formulations. These last several factors illustrate the great need for comprehensive global monitoring of all radiatively significant climate forcings and feedbacks [Hansen et al., 1990].

Acknowledgements. We are indebted to Christina Koizumi for typesetting, Jeff Jonas for color work, the NASA Climate Program Office and EPA for support of climate model development, and John Hoffman of the EPA Global Change Program for purchase of an advanced work station for GCM simulations.

References

Angell, J. K., Impact of El Nino on the delineation of tropospheric cooling due to volcanic eruptions, *J. Geophys. Res.*, *93*, 3697-3704, 1988.

Angell, J. K., and J. Korshover, Comparison of tropospheric temperatures following Agung and El Chichon volcanic eruptions, *Mon. Wea. Rev.*, *112*, 1457-1463, 1984.

Charlson, R. J., et al., Climate forcing by anthropogenic aerosols, *Science*, (in press).

Franklin, B., Meteorological imaginations and conjectures. *Manchester Literary and Philosophical Society Memoirs and Proceedings*, *2*, 122, 1784.

Hansen, J. E., and A. A. Lacis, Sun and dust versus greenhouse gases, *Nature*, *346*, 713-719, 1990.

Hansen, J. E., and S. Lebedeff, Global trends of measured surface air temperature, *J. Geophys. Res.*, *92*, 13,345-372, 1987.

Hansen, J. E., and S. Lebedeff, Global surface air temperature update through 1987, *Geophys. Res. Lett.*, *15*, 323-326, 1988.

Hansen, J. E., W. B. Rossow and I. Fung, The missing data on global climate change, *Issues Sci. Tech.*, *7*, 62-69, 1990.

Hansen, J. E., W. C. Wang, and A. A. Lacis, Mount Agung provides a test of a global climatic perturbation, *Science*, *199*, 1065-1068, 1978.

Hansen, J. E., et al., Climate response times: dependence on climate sensitivity & ocean mixing, *Science*, *229*, 857-59, 1985.

Hansen, J. E., et al., Global climate changes as forecast by the Goddard Institute for Space Studies three-dimensional climate model, *J. Geophys. Res.*, *93*, 9341-9364, 1988.

Hofmann, D. J., and J. M. Rosen, Sulfuric acid droplet formation and growth in the stratosphere after the 1982 eruption of El Chichon, *Science*, *222*, 325-327, 1983.

Jones, P. D., The influence of ENSO on global temperatures, *Climate Monitor*, *17*, 80-89, 1989.

Lacis, A. A., J. E. Hansen, and M. Sato, Climate forcing by stratospheric aerosols, *Geophys. Res. Lett.*, (submitted).

Lacis, A. A., D. J. Wuebbles, and J. A. Logan, Radiative forcing of climate by changes in the vertical distribution of ozone, *J. Geophys. Res.*, *95*, 9971-9981, 1990.

Mass, C., and S. Schneider, Statistical evidence on the influence of sunspots and volcanic dust on long-term temperature records, *J. Atmos. Sci.*, *34*, 1995-2004, 1977.

Robock, A., The volcanic contribution to climate change of the past 100 years, in *Greenhouse-Gas-Induced Climatic Change*, ed. M. E. Schlesinger, pp. 429-443, Elsevier, New York, 1991.

Self, S., and M. R. Rampino, The relationship between volcanic eruptions and climate change: still a conundrum? *Eos*, *69*, 74-75, 85-86, 1988.

Stommel, H., and E. Stommel, *Volcano Weather*, 177 pp., Seven Seas Press, Newport, R.I., 1983.

J. Hansen, A. Lacis, R. Ruedy and M. Sato, Goddard Institute for Space Studies, 2880 Broadway, New York, NY 10025.

(Received: October 3, 1991;
accepted: October 17, 1991)